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OPTICAL DEPTH, SIZE DISTRIBUTION AND FLUX OF DUST FROM OWENS LAKE, CALIFORNIA

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ABSTRACT

A joint United States/Russian/French collaborative experiment was undertaken in March 1993 and March 1996. Projects LODE I and II (Lake Owens Dust Experiments) took place on the anthropogenically desertified playa (dry lakebed) and surrounding regions of Owens Lake, in east-central California. One of the five parts of Project LODE was to determine relationships between optical depth and flux of dust emitted from the dry lake. Project LODE II included subsequent dust plume measurements and size distributions obtained through April 1996, to further refine the flux measurements for distinct mineral aerosol source regions at Owens Lake.

Size distributions of dust aerosol were determined and aerosol optical depths were calculated from sunphotometer solar extinction measurements taken downwind in plumes coming from the emissive areas of Owens Lake. This source was visually observed for 10 measured dust storms. The plume mass was calculated to be $1{\cdot}5 \times 10^9$ g using ground-based measurements and $\geq 1{\cdot}6 \times 10^9$ g from satellite data. Project LODE II results were found to be consistent with LODE I results for the south end of the playa, but flux values were found to be reduced for the northeastern portion of the playa by comparison. Vertical flux values estimated by sunphotometry were found to be consistent with values estimated via a micrometeorological method. Copyright © 1999 John Wiley & Sons, Ltd.

KEY WORDS: dust; flux; emissions; optical depth; Owens Lake, CA

INTRODUCTION

The global aerosol, which consists of a large contribution from arid region dusts, is currently of high interest because of its direct cooling effect. Solar radiation back-scattered by aerosols is believed to be of sufficient magnitude to offset global trace gas warming (Charlson *et al.*, 1987). Therefore, there is a need to assess the global aerosol and to determine those factors that affect its magnitude. Satellites will be a valuable tool for making such assessments. However, to achieve a reliable observation methodology, studies should be performed to quantify the amounts of dusts lifted from various surfaces. Studies also needed to be done of the change of dusts as they become entrained into the major global circulation aerosol transport streams that travel great distances around the globe. The present study quantifies the dust flux for some dust events from individual source areas on Owens Lake.

In the late 1800s, Owens Lake, the remnant of Pleistocene Lake Owens, a large (280 km^2) saline and alkaline body of water, was fed by the Owens River and local streams draining the Sierra Nevada. As a result

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of block faulting, the Owens River had terminated (no surface outflow) at the lake for at least the past 2000 years, and possibly longer (Gale, 1914; Bradbury, 1997). The lake was thus a sink for the evaporative concentration of salts brought in by both surface and ground water. As a result of a combination of climatic variation and water diversions in the Owens River valley to the north of the lake, the lake surface receded to its current essentially dry status by 1929 (Reinking *et al.*, 1975). Precipitation of a salt crust from the desiccation of Owens Lake was observed in 1921 (Black, 1956). If the Owens River had not been diverted into the Los Angeles aqueduct, the lake probably would not be dry today. The 1988 US Geological Survey datum for the lowest elevation of the lake is 1082·3 m.

When the lake evaporated, dissolved salts were concentrated by evaporation and ultimately precipitated out of solution. Although the thickest evaporite deposits in the centre of the playa have commercial value, much of the exposed lakebed's residual salts are not economically mineable at this time. The lake surface reflects significant variability in surficial composition in terms of geology and texture. There are exposed lacustrine clays, massive salt beds, thin salt crusts, sandsheets and dunes on the present lake surface (Cochran *et al.*, 1988).

Following atmospheric precipitation events (rain or snow) and during winter temperature minima, salts on the surface of the lakebed hydrate and expand, causing breakup of the surface (St. Amand *et al.*, 1986; Gill, 1995). When the surface temperature of the lakebed reaches 40°C in midsummer, some of the hydrated salts admixed in the surface layer lose their water of hydration and wet the surface (St. Amand *et al.*, 1987). This can form a cemented crust. Such wetting was physically observed by the first author to occur during spring 1995. During the late winter and early spring, an efflorescent, fragile crust grows from complete wetting of the surface giving it a snow-covered appearance (St. Amand *et al.*, 1987; Cahill *et al.*, 1996). During periods of moderate winds, the fluff and fine powder are quickly lofted into extensive dust storms. A third relatively hard type of crust is composed of salt, silt and clay (Cahill *et al.*, 1994). This crust underlies the fragile efflorescent crust outlined above. This type of crust is broken up and degraded into a rough surface by effects of desiccation and abrasion. When dust storms continue in duration the salt–silt–clay and cemented crusts are abraded by blowing sand which provides an additional source of mineral aerosols (Cahill *et al.*, 1996). Resultant dust storms have been reported as far south as the Los Angeles basin 400 km away (St. Amand *et al.*, 1986; MacKinnon and Chavez, 1996). Locally these dust storms are noted to restrict visibility moderately to severely. Dust particles smaller than 10 μm aerodynamic diameter (PM_{10}) transported during Owens Lake storms are reported to be among the highest concentrations ever recorded –40 620 μg m^{-3} for North America (Cahill *et al.*, 1994).

As part of a multinational (United States/Russian/French) Lake Owens Dust Experiment (LODE I), studying the generation of the dust storms originating from the dry lakebed surface, optical depth measurements were taken and aerosol samples were obtained during dust storm activities from 11–25 March 1993. Optical depth measurements from the ground are of great importance to satellite imagery studies of dust, as was extensively discussed by Fraser (1993). Subsequent measurements were collected until completion of LODE II, in April 1996. The purpose of this work was to estimate total mass of dust removed during individual dust events. The dust storms were clearly identified by surface observations and satellite as originating on the dry bed (playa) of Owens Lake (MacKinnon *et al.*, 1996; Gillette *et al.*, 1996).

EXPERIMENTAL DETAILS

Aerosol optical depth in the visible region of the solar spectrum is closely related to the size distribution of aerosols. If the size distribution is known, then a direct relationship between aerosol mass concentration and aerosol optical depth can be established. Mie scattering calculations for spherical particles can then be used to relate the atmosphere volume mass to the volume extinction (i.e. per cubic metre of air). The advantage of this approach is that the cloud height need not be known, and vertical inhomogeneities will not have an effect on the optical depth; whereas, if there were vertical gaps in concentration, these would not be accounted for when using altitude to determine atmospheric columnar aerosol mass. For our application, two assumptions are involved. These are: (1) the particle size distribution within the dust cloud does not vary substantially in the vertical and horizontal directions for fine suspended particles; and (2) that our use of Mie scattering

Table I. Dust storm activity (in Pacific Standard Time) and wind direction at Owens Lake, California

	Date	Start time	End time	Wind direction
LODE I	11 March 1993	05:30	16:30	North
	17 March 1993	01:20	19:40	North
	18 March 1993	08:30	14:30	North
LODE II	10 January 1996	09:00	18:45	North
	25 January 1996	07:15	17:10	Variable to north
	5 March 1996	05:00	05:00 (next day)	South
	11 March 1996	13:33	16:46	West
	16 March 1996	09:00	16:00	North
	22 March 1996	11:00	17:30	South
	23 March 1996	06:00	17:50	North to variable
	18 April 1996	10:55	15:30	North

calculations for spherical particles is a reasonable approximation for soil-derived particles that are not spherical but irregular. The uncertainties arising from these assumptions are believed to be about the same order of magnitude as the uncertainties in our estimation of the cloud dimension, and average duration and wind speeds of a wind storm. Justifications of the above two assumptions are given.

1. Size distributions do not change for fine particles in the plume because for our definition of fine particle, the sedimentation velocity is a small fraction (0·1) or smaller of the root-mean-square vertical velocity. These particles are carried in the air in a similar way to gases and may be diluted in concentration but will not be changed in relative size distribution.
2. The assumption of spherical particles is an approximation of the actual shapes of the soil particles. Although they are not truly spherical, neither are they plate-like, cubic, ellipsoid or other well defined shape. They are more like spheres than another shape even though they are rough.

Measurement Programme

Times of dust storm activity and wind directions for LODE I and II are summarized in Table I. The two storms on 17 March and 18 March 1993 measured during LODE I were small storms and dust fluxes were inconclusive. They are not further discussed in this paper. Proper measurement of dust plumes required that the measurements transect the Owens River valley roughly perpendicular to the wind direction in the dusty air downwind from the source. Figure 1 shows the experimental locations and regions that are discussed in this paper.

For LODE I, the existence of a road south of the lake (State Highway 190) that runs NE–SW furnished our transect line but restricted our measurements to north winds only. Although there were several small, short-duration dust events caused by mountain-valley winds, local heating and other mesoscale processes (Cahill *et al.*, 1994) and a few large dust events for south winds, measurements collected for LODE I were only taken during three days of dust storm activity for north winds. The first dust storm began on 11 March 1993, at 05:30 Pacific Standard Time (PST). By 10:30 PST wind direction south of the lake shifted so that the dust was accumulating against the Sierra Nevada at the town of Olancha, and by 11:30 the dust plume was overtopping Olancha Peak at an altitude of 3695 m (MacKinnon *et al.*, 1996). Despite this wind shift south of the lake, winds over the lake and over Highway 190 were from the north during the duration of the storm.

For LODE II, in addition to Highway 190, a road located to the east of the lake shoreline (State Highway 136) provided a transect for measurements of winds coming from the west and south. A road located on the lakebed on the eastern side of the lake (Sulphate Well Road) provided another transect line for measurement of winds coming from the north.

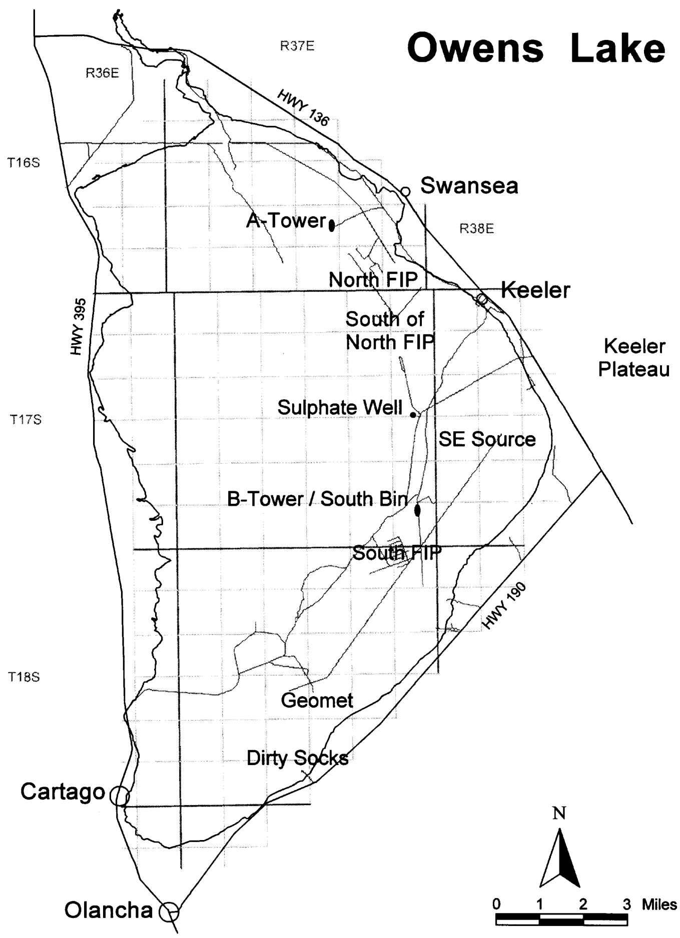

Figure 1. Map of Owens Lake in Inyo County, California. Individual source areas are labelled

Optical depths were determined using two sunphotometers custom fabricated by the National Oceanic and Atmospheric Administration (NOAA), J302 and J324. Each instrument was fitted with two filters that measured direct-beam solar irradiance at wavelengths of 380 and 500 nm. Measurements were collected along a NE–SW line approximately perpendicular to the dust plume being studied for measurements collected along Highway 190, and along a NW–SE line approximately perpendicular to the dust plume being studied for measurements collected along Highway 136. For LODE II, additional measurements were collected on Sulphate Well Road using a E–W transect line. These lines transect the shorelines of Owens Lake. For the north winds, measurements were taken along the northern edge of State Highway 190, at the State Highway mile markers, for a sampling interval of 0·5 to 1 mile. For LODE II, measurements were taken for south and west winds along the western edge of State Highway 136 at 0·5 to 1 mile intervals and along Sulphate Well Road at 0·1 mile intervals. The measurement interval was determined first by the width of the dust plume and secondly by the amount of dust. For example, fewer measurements were taken outside the dust plume than were taken within the plume boundaries. For LODE I, stratospheric dust levels were removed from the optical depth measurements. For LODE II, background measurements taken upwind of the plume were subtracted from the optical depth measurements to account for both regional and stratospheric levels.

Optical depths

Measuring the intensity of the direct solar beam, the reduction in intensity is related to optical depth in the following manner:

$$I = I_0 \ e^{-\int_0^z \xi \ dz} \tag{1}$$

where I is intensity measured, I_0 is the calibration constant determined by the Langley plot method and ξ is the extinction coefficient, determined by size distribution and complex index of refraction. Equation 1 may be rewritten as Equation 2 by using Equation 3:

$$\frac{I}{I_0} = e^{-\tau_{tot}} \tag{2}$$

Aerosol optical depth τ is defined by Beer's Law (Dutton and DeLuisi, 1983):

$$\tau = \int_0^\infty \xi(z) \ dz \tag{3}$$

where z is height. Optical depth may be used to calculate the total vertical aerosol mass assuming that the optical aerosol characteristics within the plume are constant. Several solar wavelengths may be chosen such that the only known significant attenuators are air molecules, ozone and aerosols.

Sunphotometers have been used extensively to measure solar extinction (e.g. Flowers *et al.*, 1969; Shaw, 1979; Esposito *et al.*, 1996). A summary of the performance, accuracy and precision of various sunphotometers for measuring tropospheric optical thickness, τ_{tr} (λ), was given by Dutton and DeLuisi (1982). The hand-operated two-filter sunphotometers used for this experiment measured solar extinction at two wavelengths (380 and 500 nm) and are similar to those discussed by Dutton and DeLuisi (1982). Tropospheric optical thickness can be measured using:

$$\tau_{tr} = \tau_{tot} - \tau_{st} = \int_0^h \xi \ dz \tag{4}$$

where h is the height of the top of the troposphere. Tropospheric optical thickness, τ_{tr}, was calculated by subtracting stratospheric optical thickness, τ_{st}, from total optical thickness, τ_{tot}, measured at the surface after adjustments had been made for effects of ozone and Rayleigh scattering. The value τ_{tot} is found by interpolating a least-squares fit to sunphotometer observations for wavelengths 380 and 500 nm. Time was carefully calibrated and noted for each observation so that solar angle could be used to determine the path length through the atmosphere. Estimates of stratospheric optical thickness, τ_{st}, were obtained for LODE I using Stratospheric Aerosol and Gas Experiment (SAGE) II values (Table II) (L. Thomason, NASA-Langley Research Center, Atmospheric Sciences Division, personal communication, 10 March 1994). These τ_{st} estimates were from SAGE II measured values of optical depth at 385 and 525 nm. We must note that the upper troposphere was significantly loaded with aerosols during this time period, and the error estimate obtained for the τ_{st} could be as high as ± 30 per cent. However, because these τ_{st} values are at least two orders of magnitude smaller than the calculated τ_{tr} values during LODE I owing to Owens Lake dust, the potential τ_{st} error would not significantly impact the results of this study.

There was a large gap between the time of ground sunphotometer measurements and the closest pass of SAGE II observations through the mid-latitudes. The large differences between the more southerly data and

Table II. Stratospheric optical depths for 16 and 17 March 1993 near California

Latitude (deg)	Longitude (deg)	Approximate tropopause (km)	τ_{st} ($\lambda = 0.385$)	τ_{st} ($\lambda = 0.525$)
32·6	255·1	13	0·0399	0·0416
33·1	230·9	15	0·0390	0·0447
40·3	277·3	11	0·0596*	0·0696*
40·8	253·1	12	0·0494	0·0600
41·3	229·1	11	0·0479*	0·0583*
39·4	237·7	13	0·0494	0·0521
38·9	213·5	10	0·0477*	0·0522
33·3	283·6	12	0·0376	0·0454
32·8	259·4	13	0·0500	0·0541
32·3	235·3	15	0·0254	0·0331

* Values extrapolated 1–2 km due to high opacity

the northern data reflect the location of the jet stream and, poleward of the jet, diabatically driven subsidence adds to the stratospheric opacity considerably during the winter and early spring. The values $\tau_{st} = 0{\cdot}05$ for λ_{500} and $\tau_{st} = 0{\cdot}04$ for λ_{380} were chosen to reflect this gradient. Again, however, $\tau_{st} \ll \tau_{tr}$, so that the uncertainties are not a source of serious error in our measurements. For LODE II, the same formulae were used except that regional measurements, τ_{reg}, taken upwind outside of all dust plume activity, were substituted for τ_{st}. Dust can be channelled for days from N–S and S–N in the Owens Valley owing to the effect of surrounding mountain walls (vertical relief is *c.* 1·5–3 km) to the east and west. Because of the high incidence of dust storm activity we believe that τ_{reg} would provide a more accurate measurement of combined regional and stratospheric dust.

Error analysis of sunphotometer observations

Possible sources of significant measurement errors are instrument calibration, contributions from attenuators other than aerosols, and atmospheric inhomogeneities on time and space scales that are less than the temporal and spatial resolution of the observations. It is believed that the combined random and bias errors resulting from non-aerosol attenuators would be a small and probably an insignificant percentage of the very large optical depths reported here as a result of the dust storm. Similarly, ideal calibration of sunphotometers should only produce an error of a few per cent of a measured 0·10 aerosol optical depth (Dutton and DeLuisi, 1982). The sunphotometers were calibrated before and after each LODE experiment. Measurement errors from all sources are estimated to be less than 10 per cent of the values reported here.

Calibration

Because sunphotometers measure only an instantaneous relative intensity, individual measurements were referenced to pre- and post-determined calibrations. The calibration process is critical to the accuracy when determining optical depths. Sunphotometers were calibrated by the well known Langley plot method (Box, 1981; Schmid and Wehrli, 1995). The accuracy of such calibrations is affected by the changing atmospheric conditions during the calibration process and by the stability of the instrument between the times of calibration.

Size distribution of dust plume

For LODE I, a filtration system and a high-volume inertial impactor were used to obtain aerosol samples for size distribution analysis. These samples were obtained at a site downwind of Owens Lake at an impactor intake height of approximately 3 m. Although the aerosol sampling location was limited by availability of power, we nevertheless got as near as possible to the anticipated plume centre. The location was actually 1 km south of the sunphotometer sampling line and plume centre. This location, 8 km downwind of the dust source, is close enough to plume centre that it should be representative of dust for the well mixed plume. The high-

Table III. Size distributions statistics for Owens Lake

Date*	Location	Sampler	VGMD	sg
6/19/95	Keeler	Andersen 2	2·7	3·4
09/15/95	Keeler	Andersen 2	2·3	4·2
10/4/95	Keeler	Andersen 2	2·9	2·2
10/21/95	Keeler	Andersen 2	3·5	2·0
10/26/95	Keeler	Andersen 2	3·9	1·6
03/22/96	Keeler Plateau	Andersen 1	3·3	2·1
03/22/96	Keeler Plateau	Grimm	3·6	1·7
03/11/93	Geomet/Dirty Socks	Andersen 1	3·9	1·9
03/11/96	Geomet/Dirty Socks	Grimm	4·0	2·1
03/16/96	North Fip	Andersen 1	3·2	2·1
03/16/96	North Fip	Grimm	4·4	1·7
3/22/96	S. of North Fip	Andersen 1	3·7	1·8
3/22/96	SE Source Area	Grimm	3·6	1·8

VGMD is volumetric geometric mean diameter; *sg* is geometric standard deviation
* Month/date/year.

volume inertial impactor used was the Andersen 2000 model (hereafter Andersen 1) designed to sample air at a rate of 20 $ft^3\ min^{-1}$ ($0{\cdot}57\ m^3\ min^{-1}$). The impactor consisted of four impaction plates and a backup filter. The impaction plate filters consisted of Andersen High Volume Reeve Angel Glass Fibre perforated collection discs. As these filters are designed to prevent particle bounce, they were not greased. Nominal particle separations are as follows: $> 7{\cdot}0\ \mu m$, $3{\cdot}3$–$7{\cdot}0\ \mu m$, $2{\cdot}0$–$3{\cdot}3\ \mu m$, $1{\cdot}1$–$2{\cdot}0\ \mu m$ and $<1{\cdot}1\ \mu m$ (on the backup filter). The largest particle size of the sampler was poorly defined, and we arbitrarily assigned it a value of 20 μm. The flow rate was measured with a laminar flow meter, and flow was maintained using a turbine-type pump. The intake was at a height of 1 m.

To determine more effectively the effect of size distribution and resultant extinction coefficients for Owens Lake dust from various source areas, we carried out an intensive size distribution and concurrent sunphotometer measurement experiment at Owens Lake during March 1996 (LODE II). During this experiment more than 69 size distributions were collected. These size distributions were subsequently compared to those previously collected.

In addition to Andersen 1, a second instrument was utilized for LODE II –the Grimm Model 1·105 Dust Monitor (hereafter Grimm). Air is drawn into the instrument through a nozzle set to sample air isokinetically at the ambient wind speed. The instrument classifies the particle size based on the amount of light scattered 90° from a beam produced by a laser diode. The scattered light signal is collected on a photodiode detector, amplified, and analysed by an eight-channel pulse-height analyser. Optics of the sampler were protected by sheath air produced by filtering of the ambient air. The multichannel size distribution had the following size ranges: 0·5–1·0, 1·0–2·0, 2·0–3·5, 3·5–5·0, 5·0–7·5, 7·5–10 and 10–15 μm. Particle sizes were calibrated with polystyrene latex particles. After the particles leave the optical sampling cell they are collected on a filter that has been preweighed for the particle sampling. The weight gain after sampling was compared with the total mass of particles derived from the particle size analyser. The hand-held intake was held at a height of 2 m during aerosol sampling.

Additional size distributions were included from a study conducted by the Great Basin Air Pollution Control District. These size distributions used a sampler (Andersen 2) similar to the high-volume inertial impactor discussed above. The impactor was run at $0{\cdot}57\ m^3\ min^{-1}$ onto ungreased glass fibre filter substrates. The $0{\cdot}57\ m^3\ min^{-1}$ particle size cut-points were based on experimental calibrations provided by the manufacturer. The impactor uses a sampling head that removes particles larger than a 10 μm aerodynamic cut point.

During LODE II, a propane generator source was used to power the high-volume inertial impactor, Andersen 1, so that samples could be obtained immediately downwind of the source at a distance of less than 0·5 km. The instrument was located as close to plume centre as could be ascertained. Mass distributions were obtained by weighing the impactor substrates before and after sampling.

Complex refractive index and Mie scattering

A lognormal fit was determined from the size distribution data shown in Table III. This table includes volumetric geometric mean diameter (VGMD) and geometric standard deviation. The geometric standard deviation (sg) is the ratio of the diameter below which 84·1 per cent (which is one standard deviation from the mean) of the particles lie to the median diameter, $\overline{D}_p$. For monodisperse aerosol $sg = 1$. In addition, 67 per cent of all particles lie in the range from $\overline{D}_p / sg$ to $\overline{D}_p\, sg$, and 95 per cent of all particles lie in the range from $\overline{D}_p / 2sg$ to $2\,\overline{D}_p\, sg$ (Seinfeld, 1986). Variations in calculated optical characteristics of dust aerosols are related to both uncertainties in the particle size distribution and uncertainties in prescribing a value for the complex refractive index m:

$$m = n - i æ \tag{5}$$

where the real part of m is n, the ratio of light velocity in a vacuum to light velocity in a substance; and the imaginary part æ describes absorption of light by the substance which is related to the absorption coefficient k, used in the Bouguer–Lambert law for a bulk material:

$$æ = k\lambda/4\pi \tag{6}$$

where λ is the wavelength of the incident light. It was shown in Sokolik *et al*. (1993) that in performing an analysis of optical constants of dust aerosol there is appreciable disagreement regarding the values of æ(λ). Consequently, in order to develop a model of optical constants of dust aerosols with possible peculiarities of a given geographical area, a detailed investigation is needed of the complex refractive index which has been correlated to simultaneous studies of chemical composition. This was beyond the scope of this study. For the mineral aerosol components at Owens Lake, the complex refractive index for λ_{380} and λ_{500} was assumed for the LODE experiments to be, based on other measurements of desert dust (Sokolik *et al.*, 1993):

$$m = 1 \cdot 54 - 0 \cdot 004i$$

To justify the use of the above index of refraction, we made a sensitivity study on the calculated extinction coefficient of the index of refraction. To establish a range of values for the real part of the index of refraction, we used the maximum and minimum of n for the following minerals that are found at Owens Lake: halite, anhydrite, gypsum, calcite, smectites, illites, kaolinites, quartz and plagioclase feldspars. That range was 1·5 to 1·65. For the imaginary index of refraction, our range was from 0 to −0·01. For combinations of the four real parts (1·5, 1·54, 1·6 and 1·65) and three imaginary parts (0, −0·004 and −0·01) the extinction coefficient was calculated. The maximum value of these 12 coefficients differed by 0·5 per cent of the value for index of refraction (1·54 − 0·004i) at a wavelength of 500 nm (0·4 per cent for 380 nm) and by 0·9 per cent for the minimum value (0·6 per cent for 380 nm). Applying this analysis to our calculations, we estimate that an error of less than 1 per cent results from our choice of index of refraction

The Mie scattering programme was developed at the NOAA Surface Radiation Research Unit of the Air Resources Laboratory and used corrections for time, longitude (used to calculate solar angle), pressure and Rayleigh scattering. Particles were assumed to be spherical. The input data were the measured size distribution and total mass for each aerosol sample. Total mass was checked by integration of the size distribution and assumed particle density of 2·5 g cm^{-3}.

Relative plume scale height

Relative plume scale height was estimated using the following simplification of Equation 4:

$$\tau_{\mathrm{tot}} - \tau_{\mathrm{st}} = \xi(z = 0)h \tag{7}$$

where τ_{tot} is τ for the atmospheric column, τ_{st} is τ for the stratosphere, h is plume scale height, and ξ is extinction coefficient (cm^{-1}).

For reasons mentioned previously, τ_{reg}, the τ for regional troposphere and stratosphere, is substituted for τ_{st} in LODE II. The plume scale height, h, is a relative measure useful for comparing different storms, not for estimating actual height of the dust plume. Cloud height estimates were made using ξ ($z = 0$) calculated from size distribution data at the edge of the dust cloud for LODE I and at the centre of the plume for LODE II, downwind from the source.

Column mass

If the size distribution of dust is known for the dust plume then it becomes possible to estimate the column mass, *CM*, by use of Mie theory for spherical particle extinction calculated from a measure of the particle size distribution and a measure of the column optical depth. In projects LODE I and II, the particle size distribution was only available for the near-surface. However, for PM_{10} particles, settling velocity was much smaller than friction velocity and we did not expect large differences to occur in the size distribution with height at our sampler location. Other measurements of the 11 March 1993 dust event (MacKinnon *et al.*, 1996) and other southward-travelling dust plumes from Owens Lake (Reheis, 1997) indicate that the aerosols within the plume appear to be relatively well mixed to a distance of at least 30 km from the source. Because the current measurements were taken relatively close to the dust source (much less than the distance at which sedimentation changes the appearance and composition of the Owens Lake dust plume), these assumptions should hold. Total columnar mass (g m^{-2}) was estimated using:

$$CM = \rho\, h \tag{8}$$

where ρ is aerosol concentration at sampler location and height.

Horizontal flux of airborne mass

Once columnar mass has been determined, the flux of airborne mass is calculated using the following method. The horizontal flux of airborne particles, moving with the wind through an area perpendicular to the ground and to the wind direction and that has unit width, may be written:

$$\int_0^h cU\mathrm{d}z \approx \bar{\nu} \int_0^h c\mathrm{d}z \tag{9}$$

where c is the aerosol concentration, U is mean wind speed, and z is height. The integral on the right is equal to *CM* (defined in Equation 8) and mean speed of the column $\bar{\nu}$ is calculated using:

$$\bar{\nu} \approx \int_0^h U\mathrm{d}z \tag{10}$$

Because wind erosion occurs during periods of neutral to unstable stratification, the relationship of Kaimal and Finnigan (1994) is used to express the change of wind speed with height for moderately unstable atmospheres ($-2 \leq z/L \leq 0$);

$$\mathrm{d}U = \left[\frac{u^*}{k}\right]\left(1 + 16\left|\frac{z}{L}\right|\right)^{-1/4}\frac{\mathrm{d}z}{z} \tag{11}$$

where u^*, k and L (friction velocity, von Karman constant and Monin-Obukhov length, respectively) are constant. In our experience during LODE I and II, $-z/L$ was between 0 and 2 when wind erosion was taking place. We use Kaimal and Finnigan's (1994) approximation for $-2 < z/L < 0$ and assume that the gradient

Richardson number, *R*i, is approximately equal to z/L. *Ri* was calculated as:

$$Ri = \frac{\frac{g}{\theta}\frac{\partial\theta}{\partial z}}{\left[\frac{\partial U}{\partial z}\right]^2} \tag{12}$$

where θ is potential air temperature and g is acceleration of gravity. Using the measured values for wind speed at three levels and air temperature at two levels, L and u^* were estimated using Equations 11 and 12. Height of the layer h was calculated using Equation 7. Equations 11 and 12 were used to calculate the mean wind $\overline{\nu}$. Tests of this calculation showed that the mean speed was not sensitive to the height of the layer h. For example, decreasing the height of the dusty layer by 50 per cent changes the mean wind speed by a factor of 1·2 per cent for conditions typical of dust storms ($L = 1$ km, $u^* = 0{\cdot}4$ m s^{-1}, $h = 200$ m).

Total airborne mass through a vertical plane perpendicular to the plume

The total mass of dust was then estimated using the formula:

$$mass = \sum_n (HF_n)\Delta x_n \Delta t \tag{13}$$

where *HF* is horizontal flux for a given station, n is number of measurement stations, Δx is the distance between stations (m), and Δt is the duration of *measured* storm production (s). The resultant flux is in grams.

Vertical flux from the surface

Vertical flux, $Flux_{\mathrm{v}}$, (g m^{-2} s^{-1}) emitted from the surface was estimated using the equation:

$$Flux_{\mathrm{v}} = \sum_n (Mass) \div \text{ source area } \div \Delta t \tag{14}$$

where n is the number of source areas, and size of source areas was estimated by visual observation. We outlined all active source regions onto a base map during times of dust storm activity. Time (t) refers to the time the source was active. The observation point for source maps was located at the 8500 ft elevation level of Cerro Gordo, overlooking the entire lakebed region from the east at the crest of the White Inyo mountain range. Maps were carefully digitized using an autocad program to provide the size of the source area in square metres. Concurrent 8 mm movies (frame/minute) plus 35 mm still photography (2 h interval) were used to verify source regions.

Total storm dust mass for duration of the storm

Total storm dust mass, TSF_1, for the entire storm duration was estimated for each of the individual dust storms. The following assumption was made: 'the collected data is representative of the entire dust storm'. Using this assumption and the fact that all source areas were not concurrently active during any given storm, we calculated a percentage to represent the actual time a given dust source was active:

$$\frac{t_1}{t} = \% \tag{15}$$

where t_1 refers to the time the dust source was active, and t refers to the total time of measurement. TSF_1 was then calculated using the formula:

$$TSF_1 = (F_{\mathrm{a}})(\text{source area})(\Delta t_2)(\%) \tag{16}$$

where t_2 represents the total storm duration (s) and F_{a} is vertical flux.

Table IV. Calculated optical properties of dust at each location (mean values are in italics)

Location	Date† (sampler)	ξ / unit mass* 380 nm	ξ / unit mass* 500 nm		A‡
Keeler	06/19/95 (Andersen 2)	0·5379E-03	0·5085E-03		1·058
	09/15/95 (Andersen 2)	0·6464E-03	0·5779E-03		1·118
	10/04/95 (Andersen 2)	0·3486E-03	0·3629E-03		0·961
	10/21/95 (Andersen 2)	0·2552E-03	0·2650E-03		0·963
	10/26/95 (Andersen 2)	0·1926E-03	0·1964E-03		0·981
			0·3821E-03		*1·016*
Keeler Plateau	3/22/96 (Andersen 1)	0·2863E-03	0·2983E-03	(*centre storm*)	0·960
	3/22/96 (Grimm)	0·2174E-03	0·2226E-03	(*edge of storm*)	0·961
			0·2605E-03		*0·974*
Geomet / Dirty Socks	3/11/93 (Andersen 1)	0·2163E-03	0·2228E-03	(*edge of storm*)	0·949
	3/11/96 (Grimm)	0·1655E-03	0·1688E-03	(*Centre storm*)	0·998
			0·1958E-03		*0·974*
North Fips	3/16/96 (Andersen 1)	0·2964E-03	0·3092E-03	(*centre storm*)	0·959
	3/16/96 (Grimm)	0·1754E-03	0·1789E-03	(*edge of storm*)	0·980
			0·2441E-03		*0·970*
S of North Fips	3/22/96 (Andersen 1)	0·2196E-03	0·2256E-03		0·973
SE Source Area	3/22/96 (Grimm)	0·2262E-03	0·2326E-03		0·972

* Extinction coefficient divided by mass concentration
† Month/date/year
‡ Calculated ratio of ξ between the two warelengths

RESULTS

Using the estimated complex refractive index values, a summary of calculated optical properties of dust near the surface is given in Table IV. ξ is the calculated extinction coefficient output from the Mie scattering program for spherical particles. For LODE I, derived aerosol optical depths for 11 March 1993 are shown in Figure 2. Times for sampling are given on each subfigure. Relative aerosol mass size distributions from the Andersen 1 instrument on 11 March 1993, 16 March 1996 and 22 March 1996 are shown in Figure 3. A summary of geometric means and geometric standard deviations for 11 March 1996 from the Grimm instrument are shown in Figure 4 along with concentration of particles smaller than 10 μm. Figure 4 shows that the size distributions are quite similar in shape even though concentration fluctuates greatly from one sample to another. Storms measured in LODE II showed essentially similar relative size distributions to those of Figures 3 and 4. Flux values were found to be smaller for the northeastern portion of the playa than for the southwestern portion.

To test the representativeness of the size distribution data we used the following relationship:

$$\frac{\tau_{380}}{\tau_{500}} \approx A = \frac{\xi(\lambda = 380)}{\xi(\lambda = 500)} = 0 \cdot 949 \tag{17}$$

where A is the calculated ratio of ξ between the two wavelengths 380 and 500 nm. The approximate A value for six cross-valley sets of τ measurements obtained for 11 March 1993 was $\tau_{380}/\tau_{500} = 0{\cdot}952 \pm 0{\cdot}0018$. The approximate A value compares very well with the exact A value of 0·949 above. Therefore, the size distribution used was representative of what we were observing optically. Moreover, it suggests that the measured size distribution is reasonably representative of the entire column. Calculated values for A are included in Table IV. The virtually flat spectral extinction is a manifestation of large particle influence on spectral extinction, typical of dust storms and volcanic eruptions (DeLuisi *et al.*, 1976, 1983).

Results of vertical flux from the surface, $Flux_v$ (in g m^{-2} s^{-1}) are presented in Table V, along with the percentage of time a source was active. Total storm dust mass produced for each storm is also given in Table V.

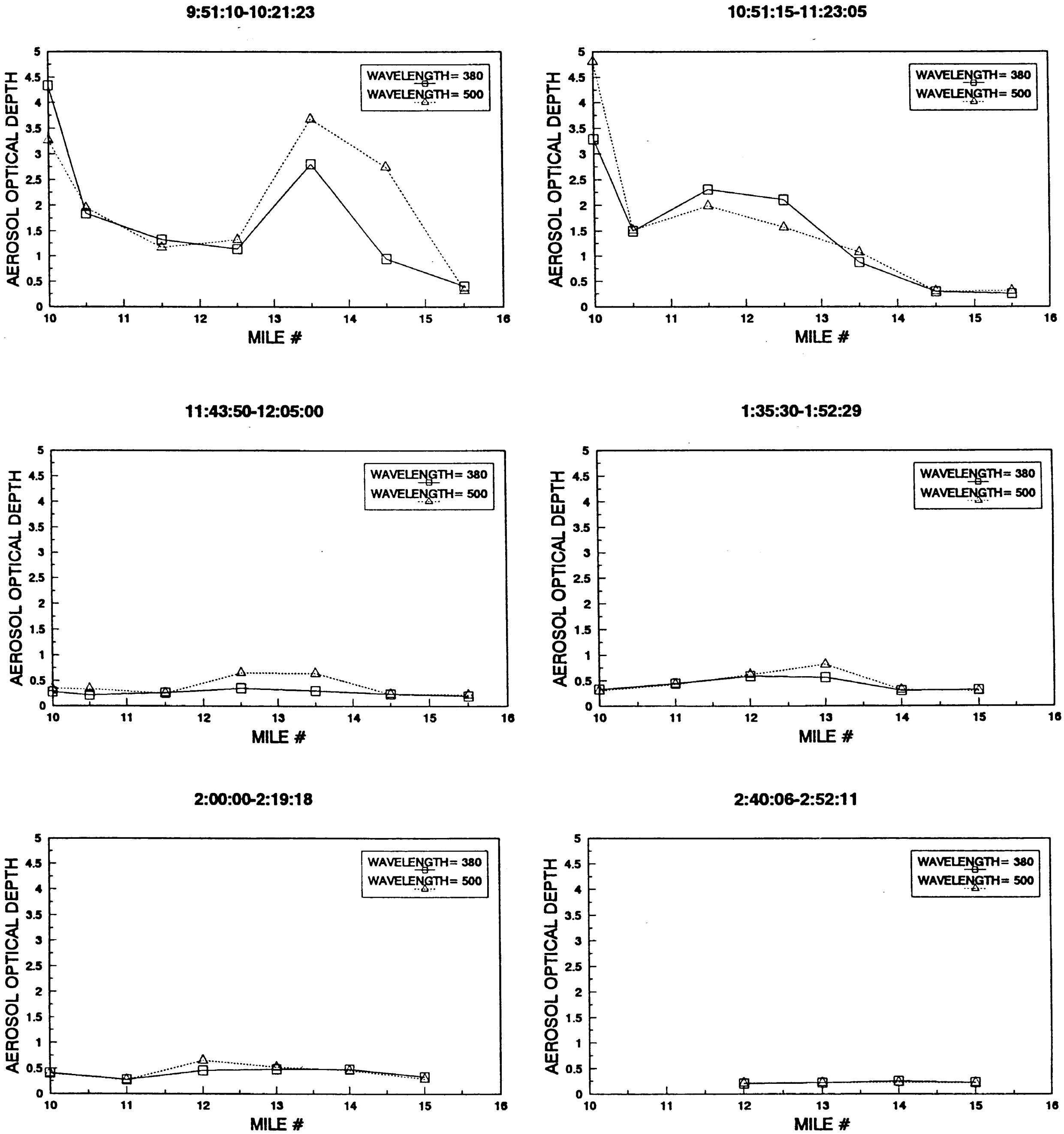

Figure 2. Aerosol optical depth versus transect distance (mile marker on Highway 190) for 11 March 1993. The plume crossed the highway between the 10 mile markers and the 16 mile markers. Time intervals for measurements are given in the subfigures

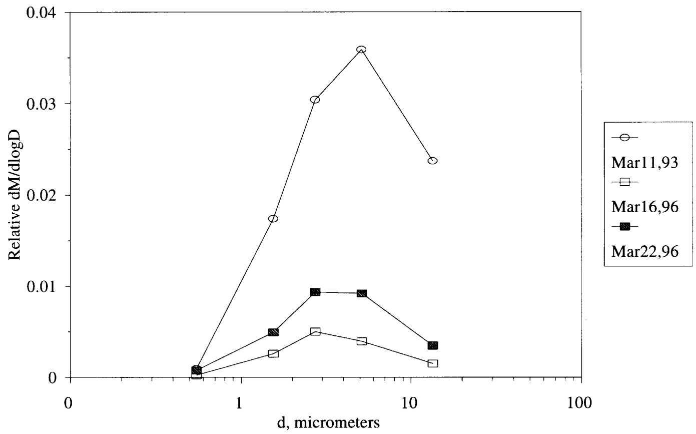

Figure 3. Relative mass size distribution for dust from the Andersen 1 instrument sampled in Owens Lake dust storms on 11 March 1993, 16 March 1996 and 22 March 1996

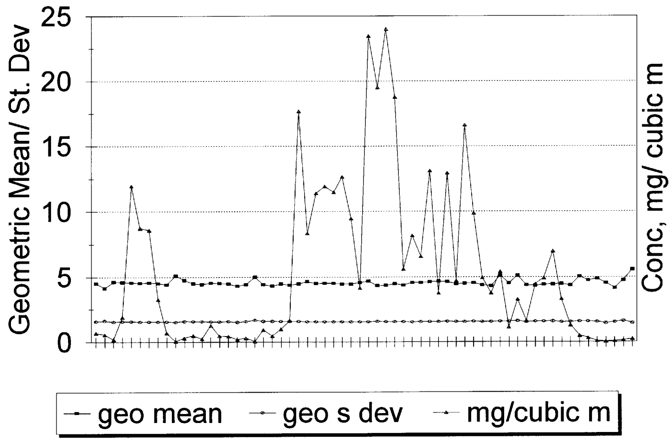

Figure 4. Geometric means, geometric standard deviations of size distributions obtained during a dust storm in the southern part of Owens Lake on 11 March 1996. Concentration of particles smaller than 10 μm is also shown. The horizontal axis is simply successive measurements that were taken at equal increments in time

DISCUSSION

Concurrent satellite measurements were available for the 11 March 1993 dust plume (MacKinnon *et al.*, 1996). For comparison we used an average (0·593) *CM* value (Run 5) to reflect dust contained within the plume later on 11 March 1993, at 14:19 PST, for the time period shortly before the end of the visible plume as seen from the ground. For a dust plume that covered an area of $2{\cdot}5 \times 10^3$ km^2 it was estimated that $1{\cdot}41 \times 10^9$ g of dust was carried in the plume. (The areal extent of $2{\cdot}5 \times 10^3$ km^2 for this event was estimated from satellite data for late on 11 March 1993; MacKinnon *et al.*, 1996) This is in roughly the same order of magnitude that was calculated for the dust plume using another method developed by Dulac *et al.* (1992) for estimating *CM*:

$$CM = c\tau \tag{18}$$

where c was the constant calculated at a value of 1·3 for desert regions at 550 μm. Using this method and an

Table V. Vertical flux, $Flux_v$, measurements, source size and estimate of total flux (by location)

Location	Measured			Total storm		
	Date*	$Flux_v$ ($gm^{-2} s^{-1}$)	Source size (m^2)	Time (s)	Active (%)	Mass (g)
South Fip	1/10/96	8·9652E-03	2·0065E + 06	35 100	81·0	5·1144E + 08
	4/18/96	7·2427E-04	3·5857E + 06	16 500	18·0	7·7131E + 06
Geomet/Dirty Socks	3/11/93	4·5594E-02	5·4130E + 06	39 600	12·1	1·1826E + 09
	3/11/93	6·7799E-01	5·4130E + 06	39 600	17·9	2·6014E + 10
	3/11/93	4·5583E-03	5·4130E + 06	39 600	21·3	2·0812E + 08
	3/11/93	2·8447E-03	5·4130E + 06	39 600	17·1	1·0427E + 08
	3/11/93	3·5567E-03	5·4130E + 06	39 600	19·4	1·4790E + 08
	3/11/93	8·2822E-04	5·4130E + 06	39 600	12·1	2·1481E + 07
	1/10/96	4·6185E-02	4·9291E + 05	35 100	100·0	7·9905E + 08
	3/05/96	1·0550E-02	2·0404E + 06	86 400	100·0	1·8599E + 09
	3/16/96	1·6147E-03	2·2111E + 06	25 200	100·0	8·9971E + 07
	4/18/96	3·4747E-04†	8·6078E + 06	16 500	74·0	3·6519E + 07
	4/18/96	1·1965E-03‡	4·5523E + 06	16 500	5·0	4·4936E + 06
	4/18/96	1·3655E-03§	3·9887E + 06	16 500	3·0	2·6961E + 06
	4/18/96	6·5256E-04¶	9·9903E + 06	16 500	18·0	1·9362E + 07
Sulphate Well (S of N Fip)	1/25/96	1·3142E-02	8·4531E + 05	35 700	14·5	5·7506E + 07
	3/05/96	7·6515E-03	9·0995E + 05	86 400	100·0	6·0159E + 08
North Fip	3/05/96	5·6028E-04	1·1251E + 06	86 400	9·0	4·6656E + 04
	3/16/96	1·3722E-03	3·4675E + 06	25 200	51·0	6·1151E + 07
	3/16/96	9·5938E-04	3·4675E + 06	25 200	49·0	4·1077E + 07
	3/22/96	3·5699E-03	1·0836E + 07	23 400	100·0	9·0519E + 08
	3/23/96	5·2923E-04	6·6531E + 06	42 600	89·0	1·3350E + 08
	4/18/96	6·7215E-04†	9·2040E + 06	16 500	80·0	8·1981E + 07
	4/18/96	2·1522E-03‡	6·0027E + 06	16 500	5·0	1·0658E + 07
	4/18/96	2·6442E-03§	4·8859E + 06	16 500	3·0	6·3951E + 06
North Fip and SE Corner	1/25/96	2·0113E-03	5·6625E + 06	35 700	2·0	8·1317E + 06
SE of South Fip	3/22/96	7·2008E-03	2·5118E + 06	23 400	96·0	4·0631E + 08
S/ K Sand Sheet	3/23/96	9·3538E-05	8·3870E + 06	42 600	100·0	3·3420E + 07

* Month/date/year
† Very early plume
‡ Early plume
§ Later plume
¶ Very late plume

optical depth, $\tau = 0{\cdot}5$, as was estimated for the minimal value contained within the plume from satellite data (MacKinnon *et al.*, 1996), we estimate that $1{\cdot}5 \times 10^9$ g of dust was carried in a plume that covered approximately $2{\cdot}5 \times 10^3$ km^2 during the late afternoon of 11 March 1993. This compares favourably with the estimate of Mackinnon *et al*. (1996) of $1{\cdot}6 \times 10^9$ g of dust in a $2{\cdot}5 \times 10^3$ km^2 plume during the same date and time. A layer of sediments averaging tens of micrometres in thickness eroded off the emissive source areas of Owens Lake could account by itself for a dust event of this magnitude (Reinking *et al.*, 1975).

The optical depths obtained reflect a dramatic increase in τ_{tr} during the 11 March dust storm and during subsequent storms that were measured. Plume definition is clearly reflected in the aerosol τ_{tr} plots that transect the valley on the southern edge of Owens Lake. Mass flux values for 1993 were found to be consistent with subsequent measurements made during 1996.

Horizontal flux values for airborne particles of sizes greater than 50 μm for the height interval 0 to 1 m q_{tot} were estimated in the source area for the dust plume measured on 11 March 1993 (Gillette *et al.*, 1997). These horizontal flux values were estimated for the times of the collection of six sunphotometer runs and were used to divide the vertical fluxes obtained in this study, given in Table V. Table VI gives the q_{tot} values along with friction velocities u_* measured in the source area of the measured plume (Gillette *et al.*, 1997).

Figure 5 shows the sunphotometry results as Owens SW (southwest). Gillette *et al.*'s (1997) LODE 1 data (Loam in Figure 5), was computed from a measured local scale vertical flux of F_a for 11 March 1993 (LODE I) from 12:00–13:30 Local Standard Time (LST) and locally measured q_{tot} giving a ratio of F_a/

Table VI. Times of erosion, horizontal mass flux, q, in source area, and friction velocity, u*, for dust storm event on 11 March 1993 at Owens Lake

Time started (h)	Time stopped (h)	q (g $m^{-1}s^{-1}$)	u_* (m s^{-1})
9·85	10·35	57·6	0·70
10·85	11·38	28·6	0·50
11·72	12·08	16·6	0·35
13·22	13·87	5·3	0·36
14·00	14·31	16·2	0·33
14·67	14·87	5·9	0·30

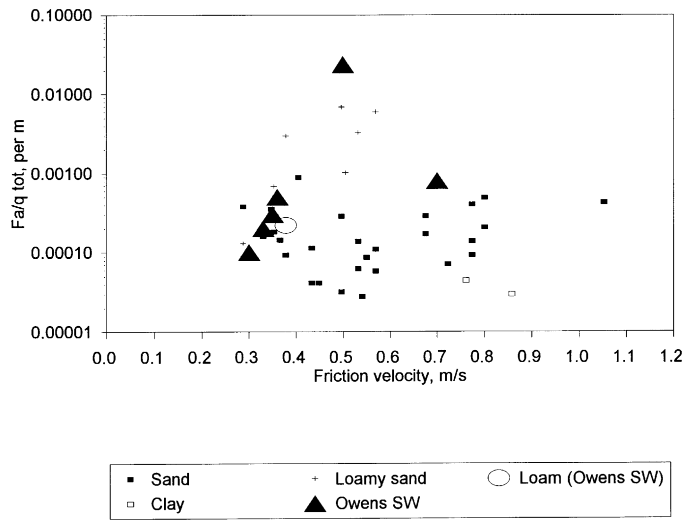

Figure 5. Ratios of F_a/q_{tot} (vertical flux of fine suspended aerosol to total horizontal mass flux) versus friction velocity for five sunphotometer flux values for 11 March 1993 (see Table V) using q_{tot} values estimated from mass flux data of Gillette *et al*. (1997) in the source area of the measured plume (solid triangles). Other values of F_a/q_{tot} are quoted from Gillette *et al*. (1997) for west Texas agricultural soils

$q_{tot} = 2{\cdot}75 \times 10^{-4}$ m^{-1}. Gillette *et al*.'s (1997) method for calculating F_a was micrometeorological using PM_{10} measurements from three towers with samplers at two heights each along with wind and temperature measurements. The F_a/q_{tot} values obtained by sunphotometry are scattered. If we except the largest value, the sunphotometry values agree with the micrometeorological value within a factor of about 3. For this highest value, observations were that dust had accumulated against the Sierra mountains which were located due west of the sampling site. That is, a secondary circulation and barrier effect of the mountains caused an increase of columnar dust mass that violated the assumptions that all dust was carried away from the sampling sites. Figure 5 also shows the F_a/q_{tot} values compared to data for wind-erodible soils of different textures in west Texas (Gillette, 1977). These results showed that the F_a/q_{tot} results for the Owens Lake playa surface material, nominally of loam texture, falls within the range of other sediment textures. The west Texas values also show a large scatter as do the sunphotometry values. The west Texas soils are agricultural soils whereas the Owens Lake surface material had higher composition of salt and carbonate. The fair agreement of the small-scale micrometeorological measurement with our large-scale values for F_a/q_{tot} from this study was satisfying.

CONCLUSIONS

We have shown that ground-based observations via sunphotometry essentially match satellite data with regard to dust plume mass. Estimates and possible errors of stratospheric and upper-tropospheric optical

depths did not appear to significantly affect sunphotometric measurements of a major dust event near its source. We believe that τ_{reg} (optical depth) taken outside of the observed plumes provides a more accurate measurement of combined regional and stratospheric to correct for dust outside of the plume than do tables of stratospheric aerosol. The aerosol measurements made using sunphotometry were reasonably representative of the entire column. Optical measurements of Owens Lake dust storms compare favourably with micrometeorological estimates, remote sensing and physical sampling of the dust. Dust flux values calculated for Lake Owens Dust Experiments I and II indicate that dust production at Owens Lake can vary among regions of the playa and/or for different times. When taken as a whole, these data help confirm that dust emanating from Owens Lake represents a substantial contribution to the regional aerosol source inventory.

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